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Connection of Low-Frequency Variations of the Earth's Pole with the North Atlantic Oscillation

V. L. Gorshkov

Pulkovo Observatory, Russian Academy of Sciences, Pulkovskoe sh. 65–1, St. Petersburg, 196140 Russia

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Abstract—The so-called Markowitz wobble (MW) is a quasi-harmonic variation of the mean pole of the Earth with a period of about 30 years and an amplitude of 0.02"–0.03". In turn, the North Atlantic Oscillation (NAO), which is characterized by large-scale phenomena in the system of atmosphere–ocean processes in this region, shows variations of some meteorological parameters in a wide frequency range. Synchronous oscillations of the pole (MW) and the NAO indices are revealed in the present study. The possibility of geophysical excitation of MW oscillations by variations of pressure fields in the North Atlantic is investigated as well.

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INTRODUCTION

The North Atlantic Oscillation (NAO) is a large-scale climatic phenomenon, which determines weather not only in the North Atlantic but also in the contiguous parts of the continents. The North Atlantic Oscillation is characterized by synchronous variations of many meteorological parameters. Another climatic formation, the North Pacific, is found symmetrically with respect to the pole in the northern part of the Pacific Ocean. These formations are sometimes considered together as an Arctic Oscillation.

Among the causes, determining the spatio-temporal NAO variations, the dynamics of large-scale interaction of atmospheric and oceanic circulations (Kushnir, 1994) and active processes on the Sun (Boberg and Lundstedt, 2003) are usually considered as the main ones.

The spectrum of pole oscillations extends from intradiurnal variations, which can be observed with modern monitoring techniques, to decadal ones, limited in this low-frequency part by the astrometric history of observations. The most powerful oscillations are nearly harmonic components of the Chandler and annual pole oscillations. A nearly linear pole trend to the American continent with a velocity of about 10 cm/yr and variations relative to this direction with an amplitude of less than a meter and a typical period of about 30 years are characteristic for the low-frequency part of the spectrum (Fig. 1).

The nature of these oscillations, which were first identified as an independent motion of the Earth's pole by Markowitz (1960), is not yet clear. Dumberry and Bloxham (2002) suggested their generation by gravitational perturbations induced by the inner core of the Earth. Sidorenkov (2002) showed that the observed average polar motion can be explained by simultaneous melting of glaciers in the Antarctic Continent and Greenland with an average loss of their mass of several grams from 1 cm^2 of the glacier surface per year. In this case, it should be assumed that various glaciers melt with a certain regularity and consistency.

The North Atlantic Oscillation, as well as El Niño, another global oceanic oscillation, is a possible "conductor" of the behavior of the Earth's rotation vector due to the angular momentum exchange with the solid envelope of the Earth and variations of its inertia tensor (Sidorenkov, 1997). Chao and Zhou (1999) examined the possibility of geophysical excitation of the pole oscillations by the NAO process in a wide spectral

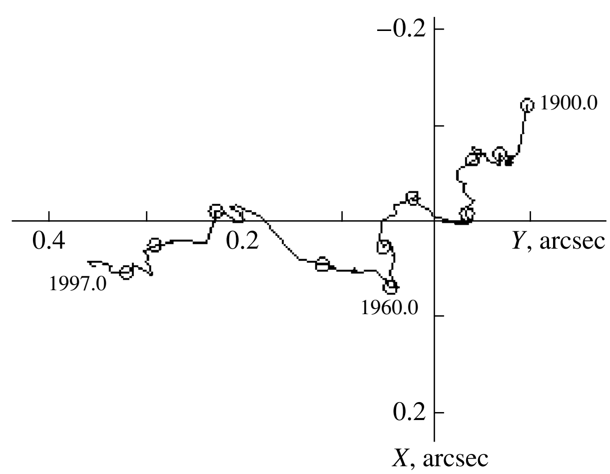

Fig. 1. Variations of the motion of the mean pole in 1900–1998 according to the EOP(IERS)C01 data.

range. They mentioned that, since the NAO activity is mainly meridional, it may be responsible for the pole oscillations, while El Niño variations, with centers of pressure contrasts distributed in latitude, should cause variations in the velocity of the Earth's rotation. However, the above authors used data starting only from 1962 and preliminarily subtracted polynomial and seasonal components from the polar motion; thus, long-period NAO manifestations were not examined.

There are two possibilities to estimate the contribution of a geophysical factor to the excitation of the dynamics of the Earth's rotation vector (only the polar motion will be considered in the present study). On the one hand, given the coordinates of the polar motion $p = x - iy$, we can estimate a possible integral excitation function $\psi = \psi_x + i\psi_y$ applying the Liuville equation (Lambeck, 1980) as follows:

$$p + (i/\sigma_c)\mathrm{d}p/\mathrm{d}t = \psi. \qquad (1)$$

Here, $\sigma_c = \omega_c\,(1 + i/2Q)$ is the complex frequency of free nutation, $\omega_c = 0.843$ is the frequency of the Chandler wobble, and $Q = 50$–180 is the Q factor of the oscillatory system at the resonance frequency.

On the other hand, if we have geophysical data in the form of pressure series and atmospheric and oceanic currents, we can estimate their contribution to the polar motion due to the mass redistribution and relative motions, in other words, due to variations of the inertia tensor $I(t)$ and angular momentum $h(t)$. In this case, the excitation function is written as follows (Wahr, 1982):

$$\psi = 1.61[h(t) + \Omega I(t)/1.44]/[\Omega(C - A)]. \qquad (2)$$

Here, Ω is the angular velocity of the Earth's rotation and C and A are the principal moments of inertia of the Earth. The numerical coefficients in Eq. (2) reflect the reaction of the elastic Earth to loading and rotational deformation. Having estimated ψ and substituted it into Eq. (1), after numerical integration, we obtain the calculated "geophysical" pole coordinates $p(x, y)$, which can be compared to the astronomic observations of the Earth's orientation parameters (EOP).

In the case of a harmonic excitation factor $\psi e^{i\sigma t}$ with a frequency σ, the forced pole oscillation is decomposed into two components coinciding with the direction of the Earth's rotation (+) and opposite to it (–) (Munk and McDonald, 1960):

$$\psi = \psi^+ e^{i\sigma t}/(1 - s) + \psi^- e^{-i\sigma t}/(1 + s). \qquad (3)$$

Here, $s = \sigma/\sigma_c$ is the ratio of the excitation frequency to the resonance frequency. Since in the case of the Markowitz wobble $s \le 0.05$, the conclusions of Munk and McDonald (1960) about the asymptotic synchronism of motions and the orientations of excitation and rotation ellipses, which become closer to each other while the excitation frequency σ decreases, are true.

The goal of the present study is to demonstrate the possibility of the geophysical excitation of long-period pole oscillations (MW) by the low-frequency component of the pressure field of the North Atlantic Oscillation.

INITIAL DATA

The Earth's orientation parameters series EOP(IERS)C01 of the International Earth Rotation Service (http://hpiers.obspm.fr/eop-pc/) were used in the study. It should be noted that this series of the pole coordinates (X, Y) begins in 1846. However, a filter removing information about the mean pole motion had been used in calculations of the polar coordinates to 1896, when the International Latitude Service (ILS) was organized. In addition, before the ILS was organized, the pole coordinates were determined with an error of more than 0.05", while the amplitudes of the Markowitz wobble under consideration are from 0.02" to 0.04".

Pressure variations between stations in the Mediterranean part of the Atlantic (for example, on the Azores (A)) and Iceland (I) are used as the NAO indices (NAOI). The NAO indices are the standardized pressure differences between these stations:

$$\mathrm{NAOI} = (P_\mathrm{A} - \bar{P}_\mathrm{A})/\sigma_\mathrm{A} - (P_\mathrm{I} - \bar{P}_\mathrm{I})/\sigma_\mathrm{I}. \qquad (4)$$

Here, $\bar{P}$ are the series-averaged pressures and σ are the mean errors of the corresponding series.

The available instrumental series of NAOI are based on barometric data and begin from the middle of the 19th century. In addition to these series, the indices "restored" by the method of principal components of the instrumental data on sea level temperature and other climatic parameters (Jones and Mann, 2004) are available starting from the 13th century. The so-called "winter" NAO indices are often used, which are obtained only for winter months, when the pressure fields in the southern (Mediterranean) and northern (Iceland) parts of the North Atlantic are most contrasting.

We used the series of monthly NAO data starting from 1824 (Jones et al., 1997) and from 1864 (Hurrell, 1995), referred to below as NAOI-1824 and NAOI-1864. The NAOI-1824 series and its amplitude spectrum are shown in Fig. 2. Although the NAO indices do not demonstrate any dominant periodicity, they contain a set of quasi-harmonic components; in particular, there are components in the low-frequency range (~24 years) that may be responsible for the MW modulation.

All series studied here have irregularities, including those associated with global climatic phenomena. Singular spectrum analysis (SSA) with the subsequent reconstruction of the significant informative components (Danilov and Zhiglyavskii, 1997) is the most adequate method to study quasi-harmonic components of series of such a nonstationary nature. The capabilities of this method for geophysical applications have been

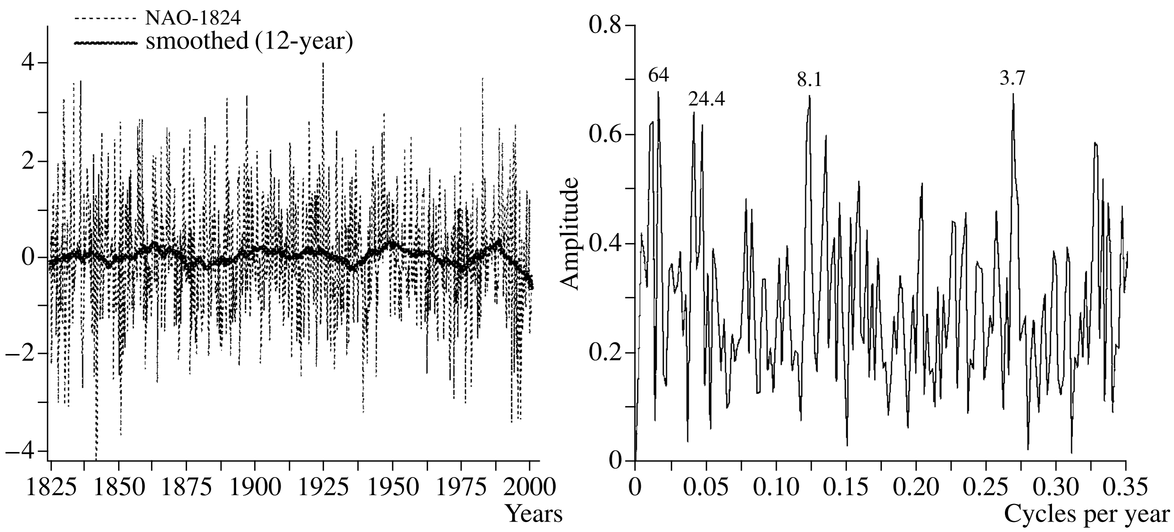

Fig. 2. Seasonal values of the NAOI-1824 index and its 12-year averaged values (left); its amplitude spectrum in the same units (right). The numbers at the spectral peaks in the left panel show the corresponding periods in years.

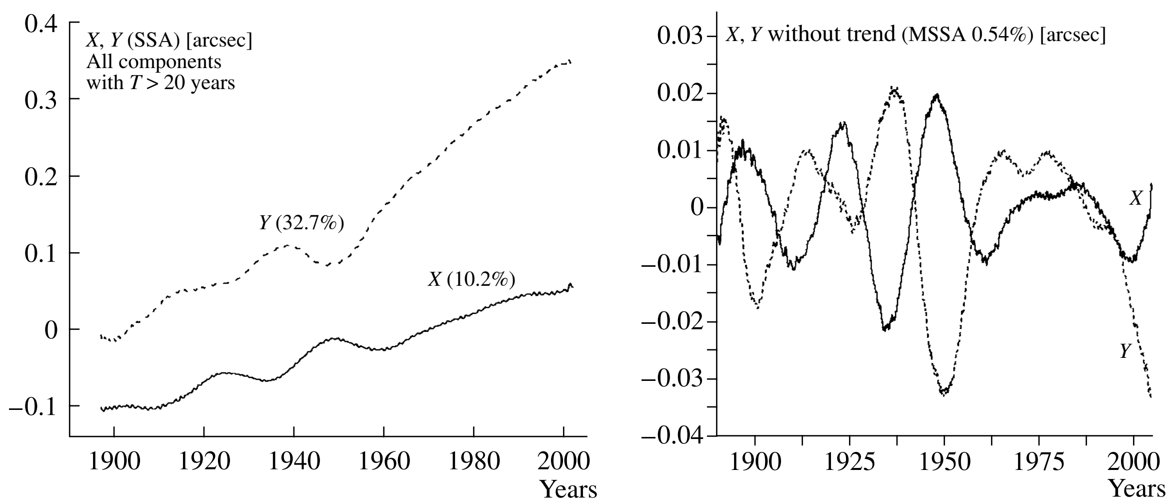

Fig. 3. All low-frequency components of the polar motion (*X* and *Y* are in arcsec), including the trend, chosen by the SSA from the EOP(IERS)C01 series from 1896 (left); only MW components extracted by the MSSA (right).

examined in detail in (Vorotkov et al., 2002). A multidimensional generalization of the SSA method (multiple SSA, MSSA) was also used for joint analysis of the series under consideration.

RESULTS OF JOINT DATA ANALYSIS

All components with frequencies $\omega \leq 0.05$ cycle per year were chosen in the pole coordinates *X* and *Y* by the MSSA (jointly for *X* and *Y*) and SSA (separately for *X* and *Y*) methods (Fig. 3). The total contribution of the MW components to the pole oscillations does not exceed 1%, since the power of this process is mainly concentrated in a nearly linear trend (38%), seasonal (15%), and Chandler (39%) components. Besides the MW, the *Y* component of the polar motion has a more powerful semi-centennial variation.

Figure 1 shows that the character of the secular motion of the pole changed in the 1950s. Therefore, the estimates of the low-frequency components were performed for two separate periods (1896–1956 and 1956–2004). The ellipses of the pole motion, corresponding to the Markowitz wobble, were estimated by the least-squares method (LSM) using SSA approxima-

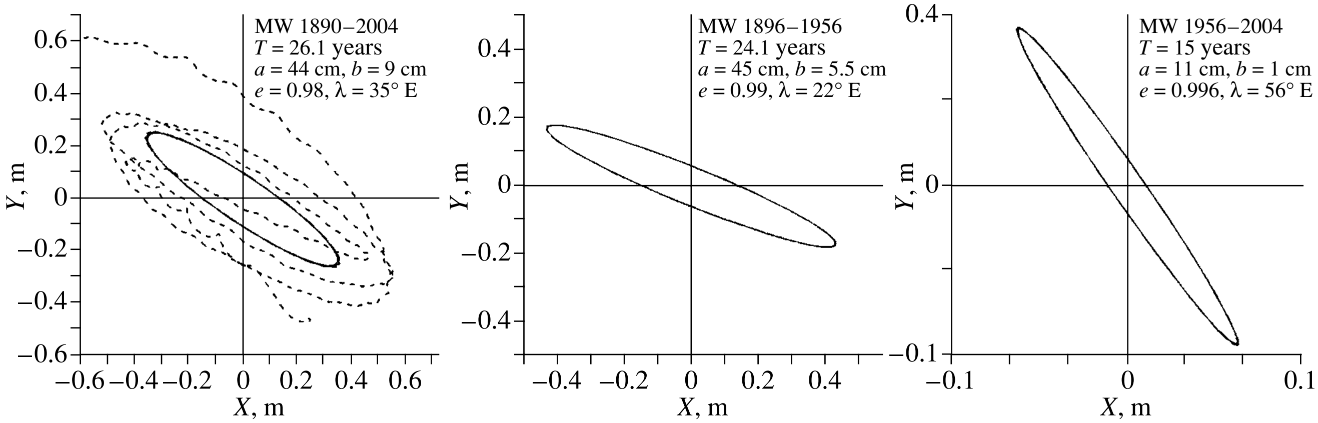

Fig. 4. MW ellipses estimated by the EOP(IERS)C01 data for 1890–2004 (left), 1896–1956 (center), and 1956–2004 (right). The dashed curves in the left figure show the results of Vondrak et al. (1998).

tions after trend removal (Fig. 4). The formal LSM errors are from 1° to 2° for the ellipse orientation, less than 0.05 years for periods, and up to 15 cm for amplitudes. However, the actual errors of all parameters should be even higher, because the SSA components used for the ellipse estimations have nonzero errors themselves.

The character of the polar motion changed nearly at the moment when the errors of the ERP determination were abruptly reduced due to implementation of a new monitoring technique. To ascertain the actual character of the low-frequency pole oscillations for the entire series, similar work was performed using the uniform EOP data obtained after reanalysis of only classic observations for 1900–1992 in the system of the ICRS catalog and the new precession–nutation model (Vondrak et al., 1998). The results are presented in Fig. 4 (by dashed curves on the left) without the LSM fitting. The resemblance of the MW characteristics obtained from these series is clear, which indicates the reality of the MW behavior regardless of the applied observation technique.

The estimated ellipse parameters are not reliable for the period from 1955 to 2004, as should be expected because of the smallness of the time interval; however, a decrease in the MW period and amplitude is obvious. The most reliable estimates were obtained for the period from 1890 to 1956. If the 1900–2004 data are used for estimates, the ellipse has the same geometric parameters but is characterized by the orientation λ = 32° E and a period of about 28 years. According to other estimates (Schuh et al., 2001; Vondrak, 1999), the MW period varies from 28 to 31 years and the amplitude is from 0.6 to 1 m.

The low-frequency NAOI components, which were extracted by a simple smoothing and by FFT filtration, are shown on the left of Fig. 5. All methods give close results in the low-frequency range. In the center of Fig. 5, there is presented a joint MSSA approximation of the standardized (and thus dimensionless) X and Y components of the Markowitz wobble and the NAOI-1824 index. The NAOI oscillations are in phase with those of the X component of the MW and in antiphase with those of the Y component. On the right of Fig. 5, the low-frequency component of the NAOI-1864 series, extracted by SSA (7%), is compared to the same estimate of the X component of MW (0.7%) and to the projection of this oscillation onto the average NAO longitude (smallest amplitude) and onto the average longitude of the major axis of the MW ellipse (naturally, the maximal amplitude).

The two last plots demonstrate that the pole oscillations with the MW frequency and the pressure variations in the North Atlantic are very synchronous; namely, an increase in the loading in the Mediterranean part of the Atlantic (NAOI > 0) shifts the mean pole in the positive direction of the X axis (towards western Europe), corresponding to variations of the inertia tensor $\boldsymbol{I}$ (mainly, the $\boldsymbol{I}_{13}$ component).

The NAO indices do not contain any information about the pressure in the region, which is necessary to calculate the excitation function. The global series of atmospheric parameters prepared by the United States National Centers for Environmental Prediction (NCEP reanalysis program, ftp://ftp.cdc.noaa.gov/pub/datasets/ncep.reanalysis/) are too short for long-period estimates. The excitation ellipse can be approximately estimated using the monthly mean pressure data at the stations in Iceland (φ = 65° N, λ = 22° W) and on the Azores (φ = 37° N, λ = 26° W), which are located almost in the centers of NAO action. These data, starting from 1865, were taken from the site (ftp://ftp.cru.uea.ac.uk/data). The mean pressures at these stations and their errors are as follows: $\bar{P}_A$ = 1021.4 ± 3.7 mbar, $\bar{P}_I$ = 1005.8 ± 7.0 mbar.

After low-frequency filtering, the pressure differences $dP = P_A - P_I$ vary within 4 mbar (within 5–6 mbar

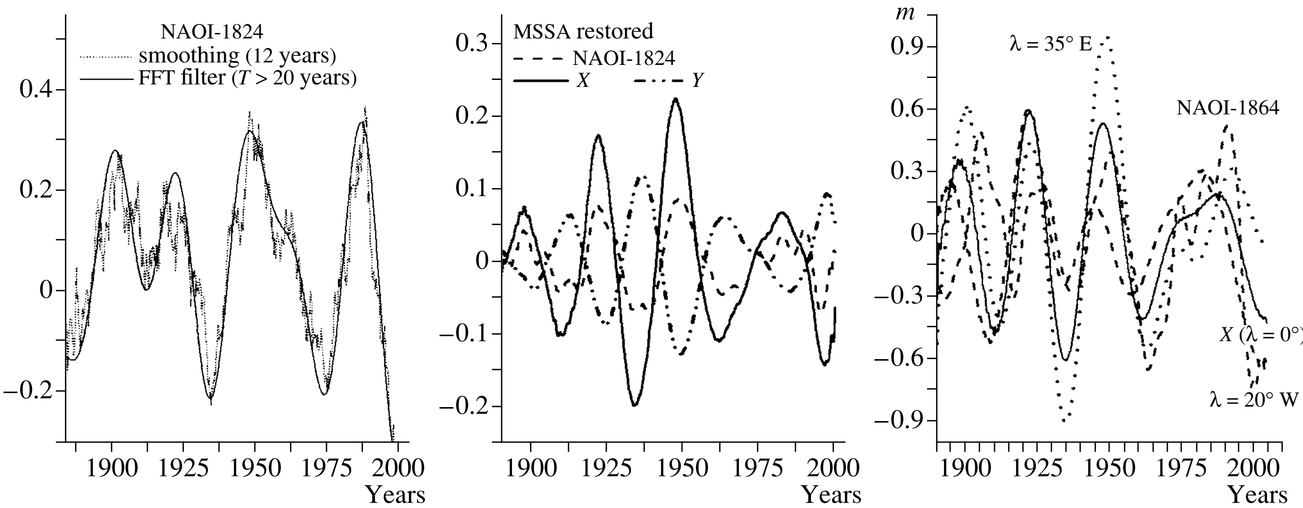

Fig. 5. The low-frequency component of the NAOI-1824 index extracted by different methods (left panel). The matched low-frequency components of the NAOI-1824 series and the pole coordinates in the MW region extracted by the MSSA from the normalized (dimensionless) data (center panel). The SSA estimates of the Markowitz wobble in the X coordinate of the pole (in meters), in the projections onto the NAO longitude ($\lambda = 20°$ W) and onto the longitude of the MW ellipse ($\lambda = 35°$ E) as compared to the same estimate for the NAOI-1864 series.

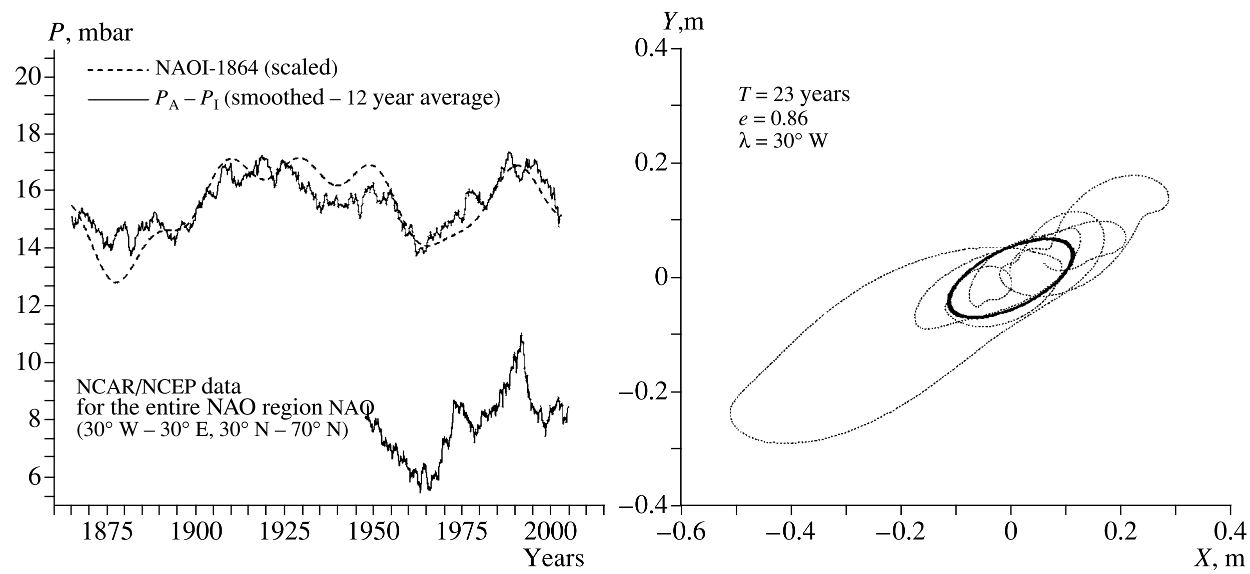

Fig. 6. Left panel: The smoothed (a 12-year moving average) difference dP (in mbar) between the stations on the Azores (P_A) and in Iceland (P_I); the scaled NAOI-1864 index is shown by the dashed line; the pressure difference (NCEP data) for the entire NAO region with the dividing parallel at $\varphi = 55°$ N is given at the bottom. Right panel: The ellipse of the excited polar motion (in meters) calculated using dP.

for NCEP data) and, as Fig. 6 (left) shows, reflect the basic regularities of the NAOI behavior. Assuming that these oscillations can be referred to the entire NAO region ($S \approx 35$ million sq. km), variations of the moment of inertia between the peaks of this oscillation should be $dI_{13} \approx dPSR^2 \cos(\varphi) \approx 10^{28}$–$10^{29}$ kg m^2, where R is the Earth's average radius, which is, according to Eq. (2), sufficient for the excitation of the pole wobble within the necessary limits (~0.02").

The low-frequency components extracted from the P_A and P_I when taking into account only the load term (second) of Eq. (2) were used to calculate the excitation function ψ, which was later used to integrate Eq. (1) in order to estimate the excited polar motion. The components of the inertia tensor in this case were calculated by the summation over only two centers ($i = A, I$) via the formulas

$$dI_{13} = K\sum_i (P_i \cos^2\varphi_i \sin\varphi_i \cos\lambda_i),$$

$$dI_{23} = K\sum_i (P_i \cos^2\varphi_i \sin\varphi_i \sin\lambda_i),$$

$$K = -R^4/[g(C-A)].$$

The component of the excitation function associated with the motion of air masses (the first term in (3)) was neglected in our study. It is relatively small owing to the multiplier σ/Ω. Because of the above assumptions, the period, orientation, and oblateness of the excited motion ellipse must be considered as preliminary results.

The ellipse of the polar motion excited by low-frequency variations of P_A and P_I is shown in Fig. 6 (right). It is naturally oriented along the line connecting the coordinates of the applied stations. Comparing Fig.4 and Fig. 6, we see that the orientation of the MW oscillation axis is about 60° to the east of the excitation axis. In addition to the fact that this estimate was obtained using only two stations, which were oriented in a certain way, there can be another explanation for the above divergence. Since NAO covers the northwestern region up to 30° E–40° east longitude and the load effect is almost completely compensated over ocean regions via the inverse barometer effect (Sidorenkov, 2002), the excitation can be strongly shifted to the eastern, continental longitudes. The additional displacement in the same direction can be due to a weaker, but slightly shifted to the east, equivalent of the NAO in the North Pacific (http://www.cpc.ncep.noaa.gov/products/). The influence of the Siberian center of atmospheric action is also possible; however, we do not have sufficiently long series for the estimation of low-frequency characteristics over this region.

CONCLUSIONS

The character of low-frequency oscillations of the pole (Markowitz wobble) changed in the 1960s; namely, the oscillations changed their orientation to a more eastern one (from 20° E to 55° E), their period decreased (from 25 to 15 years), and the amplitude of oscillations decreased strongly (from 0.025" to 0.008").

The North Atlantic Oscillation has enough power of pressure field variations in the low-frequency range to excite the Markowitz wobble in the polar motion. The estimates of the orientation of the excitation ellipse lie in the meridian of the basic direction of the NAO action (about 20° W–30° W).

The ocean reaction to the variation of the atmospheric pressure field (as an inverse barometer) can lead to the displacement of the orientation of the excitation ellipse to the direction coinciding with astronomical estimates of the orientation of the rotation ellipse for the Markowitz wobble (20° E–35° E).

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